THE GEOTHERMAL GRADIENT OF IO. G. Leone and L. Wilson, Environmental Science Department, Institute of Environmental and Natural Sciences, Lancaster University, Lancaster LA1 4YQ, U. K.

The extremely high thermal flux emitted from Io is generated by the tidal interaction with Jupiter and its other satellites Europa and Ganymede. The energy dissipation produced by this tidal interaction is very important and determines the production of internal heat which is then released from volcanic centres scattered all over the Ionian surface. The presence of volcanic centres on Io was already known since 1979 at the time of the Voyager missions but, for many years, no temperatures higher than 700 K have been measured. Only in 1986 were higher temperatures of 900 K measured from ground-based experiments [1] suggesting magmatic material of exclusively silicate composition. Recent measurements of the anomalous thermal emission from Io's volcanoes [2, 3] have revealed that magmas are erupted at temperatures as high as 1500 K and even 1800 K, temperatures too high even for magmas of basaltic composition. Therefore, these temperatures may be consistent with magmas of ultramafic composition like komatiites (see also [4]).

Clearly, thermal emissions of 1000 K are widespread and have changed noticeably our views about the Ionian crust composition. Following these observations, the issue of the Iothermal gradient has come on to the scene again. The Iothermal gradient has already been studied by [5] who calculated a value of 200 K km^{-1} based on a nominal heat flux of 100 erg cm^{-2} s^{-1} and on the assumption, widely relying on the resurfacing rates given by [6] and on a 20 km thick sulphur-rich lithosphere, that 25% of the heat flux would escape through volcanism while 75% would be conducted upward through the crust. New calculations based on a good compromise among the models published so far by many authors [5, 6, 7, 8, 9, 11] about the materials constituting the Ionian surface, on new Galileo heat flux measurements of 2.5 W m^{-2} [8], and on some widely accepted starting points including a lithosphere of mainly basaltic composition not less than 30 km thick, very high magma temperatures up to 1800 K, and a global heat flux of 2.5 W m^{-2} show that these figures are not correct. Although the model itself is very interesting and still valuable, it needs some adjustment in the respective material percentages, as a thick basaltic lithosphere of Io conducts heat very slowly and may account for 0.09 W m^{-2} (nearly equal to the 100 erg cm^{-2} s^{-1} assumed in [5]) while volcanoes are responsible for the remaining 2.41 W m^{-2} if other sources are neglected. At this point, the figures would change in such a way that the heat delivered to the surface due to conduction through the crust would account for 3.6% of the total, implying that volcanism is the main process able to maintain the heat flux observed. In this case, the lithosphere should be colder than expected in earlier studies dramatically shifting the Iotherms downwards. The 198 K Iotherm, which [5] put at a depth between 500 and 1000 m, would now lie much deeper at 4 km in a 50 km thick crust. Starting from the assumptions made in [8] which basically relies on the model made by [7] in which the resurfacing rates are considered equal to subsidence rates to explain the balance between new crust produced after volcanic eruptions and consumed crust at the base of the lithosphere (the recycling), we see that resurfacing rates higher than 15 mm a^{-1} [8] and much higher than 1 mm a^{-1} [9], even 730 mm a^{-1}, are possible to justify the observed heat flux. Furthermore, heat is conducted upwards through the lithosphere slightly faster than lithosphere subsides. Our figures are 2.7 Ma for heat rising against the 10 Ma required for a complete subduction of the surface material to the base of the lithosphere. In this way, a zero balance should not be expected because heat rises faster than descending material and a heat flow, even small, does exist. Of course, fast rising material is then redistributed on larger subsiding surfaces. The only important thing is that we can find solid SO_2 much deeper within the crust but even this fact is not crucial to the development of the volcanism since SO_2 alone does not contribute in a significant way to decrease the bulk density of rising magmas to the required level necessary to overwhelm the neutral buoyancy traps located at shallower levels [10].

Now, the question that spontaneously arises is what would a suitable geothermal gradient for Io be? Of course, this depends upon local geological settings (i.e. near to or away from volcanic centres). Given that conduction is not an effective way for heat transfer through the lithosphere and does not justify the observed heat flux, we see that heat can only be delivered upwards effectively by convective or advective processes occurring within rising magmas. The presence of a magma reservoir, whose depth may be easily estimated depending upon magma composition according to our recent model [11], raises the temperature and affects the trend of the Iotherms expected for the geological setting assumed and so the 1800 K Iotherm would not be near to the base of the lithosphere but less deeper somewhere in the crust. Estimating the Iothermal gradient in the presence of a hotspot is quite easy thanks to the reliable assumptions that can be made about magma composition, its melting temperature, and its temperature at the surface. In fact, in conduits leading to the surface, thermodynamic arguments made by [5] still hold in our model [11] and we consider magma rising nearly adiabatically from reservoirs located at deeper levels, in which case we find a very low thermal gradient ranging from 1 to 4 K km^{-1}. However, further changes to the entropy vs density diagram given in [5] should be made in consideration of the detected presence of H_2S and H_2O in Ionian magmas [12], although in small amounts. After condensation, these gases are then mixed with the pyroclastic material and recycled back into the lithosphere and, due to the low escape rate which can remove only a fraction of the material being brought to the surface by volcanism, dissociated in minor amounts to supply the hydrogen cloud detected at Io's poles and in the torus (recycled-dissociated gas ratio should be nearly 10000:1 according to [9]).

Away from volcanic centres, based on some simple assumptions and on suitable application of the equation for the one-dimensional cooling of a semi-infinite half space, our estimates for an average Iothermal gradient range from ~30 K km^{-1} to ~60 K km^{-1} depending upon lithosphere thickness ranging from 50 to 30 km, respectively. However, given the actual uncertainties in magma compositions reaching the surface and on lithosphere thickness, these values will be subject to some variations due to further improvements of the available models and new data which should be gathered in late 1999; however, the order of magnitude should not be significantly different.

References: [1] Johnson T. V. et al. (1988) *Science 242*, 1280. [2] Lopes-Gautier R. et al. (1997) *GRL 24*, 2439. [3] McEwen A. S. et al. (1997) *GRL* 24, 2443. [4] Matson D. et al. (1998) *LPSC XXIX*, n. 1650. [5] Smith B. et al. (1979) *Nature* 280, 738. [6] Johnson T. V. et al. (1979) *Nature 280*, 746. [7] O'Reilly T. C. and Davies G. F. (1981) *GRL 8*, 313. [8] Carr M. H. et al. (1998) *Icarus 135*, 146. [9] Keszthelyi, L., and McEwen, A.S. (1997) *Proceed. of the Io Conference, Flagstaff*. [10] Leone G. and Wilson L. *JGR* (manuscript in preparation). [11] Leone G. and Wilson L. (1998) *LPSC XXIX*, n. 1358. [12] Salama F. et al. (1990) *Icarus 83*, 66.